II.2-SNOW-17 HYDRO-17 SNOW MODEL

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Introduction

The SNOW-17 snow model is a snow accumulation and ablation model. It is a conceptual model in that each of the significant physical processes affecting snow accumulation and snowmelt is mathematically represented in the model.

Figure 1 [Bookmark] is a flow chart of the model showing each of the physical processes that are included.

This model has evolved from two earlier snow cover models [Anderson and Crawford (1964), Anderson (1968)]. The model described in this chapter is essentially the same as that described by Anderson (1973). There are a few minor changes which correct for some oversimplifications and errors in the earlier version of the model.

This Chapter describes the mathematical representations used in the model and explains their physical basis.

Information about how to determine model parameters is in Chapter IV.2 [Hyperlink].

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SNOW COVER ENERGY EXCHANGE

A description of the physics of snow cover energy exchange helps to understand the snow accumulation and ablation model. Of primary importance is the energy exchange at the snow-air interface. Under most conditions and especially during melt periods most of the energy exchange occurs at the snow surface. This section contains a simplified summary of energy exchange across the snow-air interface. A more complete discussion of snow cover energy exchange and heat transfer within the snow cover is given by Anderson (1976).

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Snow Cover Energy Balance Equation

The energy balance of a snow cover can be expressed as:

$$\Delta Q = Q_n + Q_e + Q_h + Q_g + Q_m \qquad (1)$$

where ΔQ is the change in the heat storage of the snow cover
Q_n is the net radiation transfer
Q_e is the latent heat transfer
Q_h is the sensible heat transfer
Q_g is the heat transfer across the snow-soil interface
Q_m is the heat transfer by mass changes (advected heat)

The units of each term in Equation 1 are energy per unit area. A useful unit in which to express these energy transfer terms is mm where a MM of energy per unit area is defined as the energy required to melt or freeze 1 MM of ice or water, respectively, at zero DEGC (approximately 8 CAL/CM2). A MM of energy per unit area will be designated by MMe to avoid confusion with a MM of length.

The change in heat storage term (ΔQ) consists of the energy used to melt the ice portion of the snow cover, freeze liquid water in the snow and change the temperature of the snow. All of these processes do not always occur during a given time interval. Water in both the liquid and solid phases can exist in a snow cover. The liquid-water is the result of melting or rainfall. The liquid-water may freeze or remain in storage. When the snow cover is isothermal, at zero DEGC and saturated (in this condition a snow cover is commonly referred to as 'ripe') the excess liquid-water becomes outflow from the bottom of the snow cover.

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Net Radiation Transfer

All bodies radiate energy. The amount and spectral distribution of the radiation is a function of the temperature of the body. In general the higher the temperature the greater the amount of radiation and the shorter the wavelength of the maximum intensity. The sun, being very hot, emits radiation at shorter wavelengths than terrestrial bodies such as clouds, trees and the snow cover. Most of the energy from the sun is in the ultra-violet, visible and near infrared portions of the electromagnetic spectrum at wavelength less than about 3 um. Most of the radiation emitted by terrestrial bodies is at wavelengths greater than 3 um. For this reason solar radiation is sometimes referred to as shortwave radiation and terrestrial radiation is referred to as longwave radiation.

The total amount of energy emitted by a terrestrial body is given by Stefan's law:

where E_t is the total emitted energy (Mme/SEC)

$$E_t = \varepsilon * \sigma * T^4 \tag{2}$$

ε is the emissivity in the longwave portion of the energy spectrum
σ is the Stefan-Boltzmann constant ($1.7*10^{-13}$ MMe/DEGK^{-4}/SEC)
T is the surface temperature (DEGK)

A body that emits the maximum amount of radiation for a given temperature (ε = 1.0) is referred to as a blackbody. Emissivity also equals absorption thus a blackbody not only emits the maximum amount of radiation but also absorbs all terrestrial radiation incident upon it. Also ε = 1.0 - r where r is reflectivity in the longwave portion of the spectrum. The reflectivity of a body to shortwave radiation is referred to as the albedo (A) of the body. Absorption of shortwave radiation is thus equal to 1.0 - A.

Snow has been found to be a nearly perfect blackbody with respect to longwave radiation, i.e., the emissivity of snow is about 1.0. Whereas snow is nearly a blackbody with respect to longwave radiation snow is highly reflective in the shortwave portion of the spectrum. This is evident to anyone who has been out in the snow on a clear and sunny day. The albedo of snow varies from nearly 0.9 for new-fallen snow to less than 0.5 for a well-aged snow cover.

The net radiation transfer for a snow cover can be expressed as:

$$Q_n = Q_i * (1.0 - A) + Q_a - \Delta t * 1.0 * \sigma * (T_o + 273)^4 \qquad (3)$$

where Q_i is the incident (incoming) solar radiation (MMe)
Q_a is the incoming longwave radiation (MMe)
Δt is the computational time interval (SEC)
T_o is the snow surface temperature (DEGC)

For a Δt of 6 hours (the basic computational time) Equation 3 can be written as:

$$Q_n = Q_i * (1.0 - A) + Q_a - 3.67 * 10^{-9} * (T_o + 273)^4 \qquad (4)$$

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Latent and Sensible Heat Transfer

Both latent and sensible heat transfer are turbulent transfer processes. Latent heat transfer involves the transfer of water vapor between the air and the snow surface. The direction in which the water vapor moves is determined by the vapor pressure gradient. If the vapor pressure is less at the snow surface than in the overlying air water vapor moves from the air to the snow and vice versa. The rate at which the water vapor is transferred depends on the turbulence of the air. The amount of turbulence in the air is related to the wind speed. Thus water vapor transfer can be expressed in an equation attributed to Dalton as:

$$V = f(u_a) * (e_a - e_o) \qquad (5)$$

where V is the water vapor transfer (MM)
$f(u_a)$ is the function of the wind speed (u_a) at a height z_a above the snow surface (MM/MB)
e_a is the vapor pressure of the air at z_a (MB)
e_o is the vapor pressure at the snow surface (MB) (assumed equal to the saturation vapor pressure at the snow

surface temperature)

The heat transfer occurs when the water vapor either reaches the snow surface and condenses which releases latent heat or leaves the snow surface by sublimation which removes heat from the snow. The amount of heat transfer is equal to the water vapor transfer times the latent of sublimation L_s (L_s = 677 CAL/GM or 8.5 MMe/MM). Thus latent heat transfer can be expressed as:

$$Q_e = L_s * V = 8.5 * f(u_a) * (e_a - e_o) \tag{6}$$

Sensible heat transfer is related to the heat content of the air. The direction of the heat transfer is determined by the air temperature gradient. Heat moves from warmer or cooler temperatures. Because the snow surface is generally colder than the overlying air especially during snowmelt heat is normally being transferred from the air to the snow cover. The rate of sensible heat transfer, similarly to water vapor transfer, depends on the turbulence of the air. Since the mechanisms of transport are analogous it is often assumed that the turbulent transfer coefficients for heat and water vapor are equal. With this assumption the ratio of Q_h/Q_e (commonly referred to as Bowen's ratio) can be expressed as:

$$\frac{Q_h}{Q_e} = \gamma * \frac{T_a - T_o}{e_a - e_o} \tag{7}$$

where T_a is the temperature of the air at z_a (DEGC)
γ is the psychrometric constant (MB/DEGC) (γ= 0.00057*P_a, where P_a is the atmospheric pressure (MB))

If one substitutes Equation 6 into Equation 7 the resulting expression for sensible heat transfer is:

$$Q_h = 8.5 * Y * f(u_a) * (T_a - T_o) \tag{8}$$

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Heat Transfer by Mass Changes

The mass balance of a snow cover can be expressed as:

$$\Delta WE_t = P_x - O_s + V + V_g \tag{9}$$

where ΔWE_t is the change in the total water-equivalent of the snow cover (MM) (The total water-equivalent is comprised of the ice, liquid-water and water vapor in the snow cover)
Px is the water-equivalent of precipitation (MM)
O_s is the liquid-water outflow from the bottom of the snow cover (MM)
V is the vapor transfer between the snow and the air (MM)
V_g is the vapor transfer between the snow and the soil (MM)

Zero DEGC is generally used as the base temperature for heat storage computations in a snow cover because the snow must warm up to zero DEGC before it can melt. If the mean temperature of snow cover is below zero DEGC a heat deficit exists. Additional heat must be added to the snow cover to raise the temperature to zero DEGC before additional melt water can be generated.

If the temperature of the snow cover outflow is assumed to be zero DEGC and the heat content of the transferred vapor is assumed negligible then only the heat transferred by precipitation need be considered. The quantity of heat transferred to the snow cover by precipitation is dependent on the amount, temperature and specific heat of the precipitation. The wet-bulb temperature should be a good approximation of the temperature of precipitation because of the analogy between falling precipitation and a ventilated wet-bulb thermometer. Thus the advected heat transfer due to mass changes can be expressed as:

$$Q_m = \frac{c}{80} * P_x * T_w \tag{10}$$

where T_w is the wet-bulb temperature (DEGC)
c is the specific heat (CAL/GM/DEGC) (for snowfall c equals c_i, the specific heat of ice, c_i = 0.5; for rainfall c equals c_w, the specific heat of water, c_w = 1.0)

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Snow Cover Energy Balance

If one substitutes Equations 3, 6, 8 and 10 into Equation 1 the energy balance of a snow cover can be written as:

$$\begin{aligned}\Delta Q = {} & Q_i * (1.0 - A) + Q_a - \Delta t * 1.0 * \sigma * (T_o + 273)^4 + \\ & 8.5 * f(u_a) * [(e_a - e_o) + Y * (T_a - T_o)] + \\ & \frac{c}{80} * P_x * T_w + Q_g\end{aligned} \tag{11}$$

Normally, in order to use Equation 11, Q_i, A, Q_a, u_a, e_a, T_a, P_x and T_w are measured or estimated and T_o, ΔQ and Q_g are unknowns. Expressions can be formulated for Q_g and ΔQ so that the energy balance equation can be solved. These expressions and the solution technique are relatively complex. However there is one very important case for which a solution to Equation 1 can be immediately written. This is the case when an isothermal snow cover is melting. In this case T_o = zero DEGC, Q_g is negligible compared to energy exchange at the snow surface and Q becomes equal to the amount of melt. Any precipitation which occurs under these conditions would most likely be rain. The amount of snowmelt, M (MM), during a 6 hour period under such conditions is:

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COMPUTATION OF SNOWMELT IN THE MODEL

Energy balance computations of snowmelt require that good measurements or estimates of incoming solar radiation, albedo, incoming longwave radiation, wind speed, vapor pressure of the air, air temperature and precipitation are available. In the United States reasonably good estimates of mean areal precipitation and air temperature can generally be determined from point measurements. This statement cannot be made in regard to the other variables because of a scarcity of point measurements (especially in real-time) and difficulties in extrapolating many of the available observations from the measurement site to the watershed where the areal estimate is needed. This does not mean that there are no basins within the United States where snowmelt could be computed by using the energy balance equations. However the number of places where an energy balance could be applied are so few that another method of estimating snowmelt must be used as the primary snowmelt computational procedure. An energy balance method is still needed for use in those areas where sufficient data are available or where the benefits of the improved forecast accuracy that can be attained by using an energy balance procedure exceed the costs of obtaining the required data.

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Use of Air Temperature as the Index to Melt

The most commonly used index for computing snowmelt is air temperature. There are two major reasons for using air temperature. First air temperature data are normally readily available from both climatological and operational hydrometeorological networks. Second it has been shown in many studies that air temperature is probably the best single index to areal snow cover energy exchange.

Other variables have been used as indices to snow cover energy exchange. Net radiation is a good index to energy exchange at a point. However net radiation is highly dependent on the surface over which it is measured. Thus it would be extremely difficult, in most cases, to extrapolate point measurements of net radiation to determine an areal estimate. In addition net radiation is measured at only a few research sites in the United States. Incoming solar radiation, vapor pressure of the air and wind speed have also been used as indices to snowmelt in conjunction with air temperature or net radiation. However these variables have not proven to be good indices to snowmelt when used by themselves.

For these reasons air temperature is used as the index to snow cover energy exchange. To be more exact the snow model uses 6 hourly mean areal air temperature estimates as the index (see Chapter II.7 [Hyperlink] for the method used to compute mean areal temperature from point measurements). In the snow model air temperature is used as an index to energy exchange across the snow-air interface. This is different from the old degree-day method which uses air temperature as the index to snow cover outflow. The degree-day method does not explicitly account for those processes (the freezing of melt water due to a heat deficit and the retention and transmission of liquid-water) which cause snow outflow to differ from snowmelt.

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Snowmelt During Rain-on-Snow Periods

During rain-on-snow periods it is assumed that melt is occurring at the snow surface. Snowmelt during rain-on-snow periods is separated in the model from melt during non-rain periods because:

- o the magnitudes of the various energy transfer processes tend to be different during the two types of periods
- o the dominant energy transfer processes during rain-on-snow periods are known
- o the seasonal variation in melt rates is generally quite different for the two types of periods

If several reasonable assumptions are made the energy balance snowmelt equation (Equation 12) can be used during rain-on-snow periods. The assumptions are:

- o incoming solar radiation is negligible because overcast conditions prevail
- o incoming longwave radiation is equal to blackbody radiation at the temperature of the bottom of the cloud cover which should be close to the air temperature
- o the relative humidity is quite high (90 percent is assumed)

In this case the wet-bulb temperature is essentially the same as the air temperature. The vapor pressure is 90 percent of the saturation vapor pressure at the air temperature (e_{sat}). A simple yet accurate equation for computing e_{sat} from the air temperature is:

$$e_{sat} = 2.749 * 10^8 * \exp\left(\frac{-4278.6}{T_a + 242.8}\right) \qquad (13)$$

When these assumptions are taken into account Equation 12 can be rewritten to express snowmelt during a 6 hour rain-on-snow period as:

$$\begin{aligned} M = {} & 3.67 * 10^{-9} * (T_a + 273)^4 - 20.4 + 0.0125 * P_x * T_a + \\ & 8.5 * f(u_a) * [(0.9 * e_{sat} - 6.11) + 0.00057 * P_a * T_a] \end{aligned} \qquad (14)$$

The atmospheric pressure (P_a) used in the model is computed from the elevation of the station or area by using the 'standard-atmosphere' altitude-pressure relationship. This relationship can be approximated by the expression:

$$P_a = 1012.4 - 11.34 * E_\ell + 0.00745 * E_\ell^{2.4} \qquad (15)$$

where E_ℓ is the elevation (hundreds of meters)

The only remaining unknown in Equation 14 besides the air temperature is the wind function ($f(u_a)$). The mean wind function becomes a parameter whose value is determined during the calibration process. Thus the equation used in the snow model for computing snowmelt during a rain-on-snow period is:

$$M = 3.67*10^{-9} * (T_a + 273)^4 - 20.4 + 0.0125 * P_x * T_a + 8.5 * UADJ * ((0.9 * e_{sat} - 6.11) + 0.00057 * P_a * T_a) \tag{16}$$

where UADJ is the average wind function during rain-on-snow periods (MM/MB/6HR)

In the model the amount of rain must exceed 2.5 MM during a 6 hour period before Equation 16 is used. This makes it more likely that humid, overcast conditions occurred.

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Snowmelt During Non-Rain Periods

The energy balance equation for snowmelt is not used as a basis for estimating melt from air temperature data during non-rain periods. This is because of the wide variability in meteorological conditions that can occur. Individual periods may be characterized by various combinations of conditions from sunny to overcast, dry to humid and calm to windy. At a given location, meteorological conditions may be generally similar from day to day and year to year during the snowmelt season. However at most locations this is not the case and even when it is, typical conditions at one location vary from those at another location. Thus the generalized snowmelt equation used to estimate surface melt from air temperature during non-rain periods is empirically based.

An empirical relationship which has given good results in many investigations indicates that snowmelt is proportional to the difference between the mean air temperature and a base temperature. This relationship can be expressed as:

$$M = M_f * (T_a - MBASE) \tag{17}$$

where M_f is the proportionality factor, commonly referred to as the melt factor (MM/DEGC)
MBASE is the base temperature (DEGC)

MBASE is a parameter of the snow model.

The most commonly used base temperature is zero DEGC. Snowmelt can occur when the air temperature is below zero DEGC, e.g., on a clear, calm day when absorbed solar radiation exceeds energy losses due to longwave radiation plus sensible and latent heat transfer. However it is unlikely that significant amounts of melt will occur when the air temperature is below zero DEGC. When the air temperature is above zero DEGC, snowmelt does not always occur, e.g., on a clear night the air temperature may remain above zero DEGC and yet longwave radiation and latent heat losses will exceed any sensible heat gain. In general, when the air temperature is above zero DEGC, the energy balance is positive, indicating surface melt. Base temperatures other than zero DEGC are sometimes used to adjust for a constant bias in the air temperature data or to attempt to obtain a more nearly linear relationship between snowmelt and the temperature difference, T_a - MBASE.

Several studies have shown that the mean value of the melt factor varies seasonally. This seasonal variation in the melt factor is mainly due to an increase, as the snow season progresses, in the amount of solar radiation absorbed by the snow cover per degree of the temperature difference, T_a - MBASE. This increase in the rate that solar energy is supplied to the snow cover is partly due to the increase in the amount of incoming solar radiation and partly due to a decrease in the albedo of the snow cover. Seasonal variations in other meteorological variables like the vapor pressure of the air, wind speed and cloud cover also influence the seasonal variation in the melt factor. Since the variation in solar energy has a dominant effect on the seasonal variability of the melt factor, the seasonal change in the melt factor is more pronounced in regions where solar radiation dominates the energy transfer process during snowmelt.

In the early stages of development of the snow model the seasonal melt factor variation was specified by a table of monthly values. In the calibration process, it is advantageous to reduce such tables, whenever possible, to an equation involving only one or two parameters. Results from the Central Sierra Snow Laboratory near Donner Pass, California [Anderson (1968)] indicated that the seasonal variation in the melt factor could be represented by a sine function. This variation is shown in Figure 2 [Bookmark] and can be expressed as:

$$M_f = \frac{MFMAX + MFMIN}{2} + \sin\left(\frac{n*2\pi}{366}\right) * \frac{MFMAX - MFMIN}{2} \qquad (18)$$

where MFMAX is the maximum melt factor assumed to occur on June 21 (MM/DEGC/6HR)
MFMIN is the minimum melt factor assumed to occur on December 21 (MM/DEGC/6HR)
n is the day number beginning with March 21

MFMAX and MFMIN are model parameters.

Based on subsequent testing, this sinusoidal melt factor variation has proven to be adequate for use throughout the contiguous United States. However when the model was applied in Alaska, the preliminary results indicated that a sine curve did not adequately describe the melt factor variation. In order to determine the form of the seasonal variation in the melt factor for Alaska, melt was computed by the energy balance equation for each month during the snow season. Air temperature, dew-point, wind and solar radiation data from Fairbanks were used. Mean conditions on days when the air temperature was above zero DEGC were used. A simple adjustment factor was determined which could be applied to the sinusoidal melt factor curve in order to obtain the seasonal melt factor variation for Alaska. This adjustment factor, MF_{adj}, is:

$$MF_{adj} = \begin{cases} 0.0 & \text{for } x \le 0.48 \\ 1.0 & \text{for } x \ge 0.7 \\ \dfrac{X-0.48}{0.22} & \text{for } 0.48 < x < 0.7 \end{cases} \quad (19)$$

where x is the decimal fraction of the time between December 21 and June 21

The coefficients used in Equation 19 are based on the Fairbanks data and the application of the model to the Chena River at Fairbanks. No other basins were tested in Alaska. The adjustment factor is applied to the sinusoidal portion of the seasonal melt factor curve. Thus the melt factor for use during non-rain periods in Alaska can be expressed as:

$$M_{f(Alaska)} = (M_f - MFMIN) * MF_{adj} + MFMIN \quad (20)$$

The seasonal variation of the melt factor in Alaska is shown in Figure 2.

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OTHER SNOW MODEL COMPONENTS

The computation of snowmelt is the single most important part of a snow accumulation and ablation model. This is because the snowmelt process largely controls the rate at which water is released from the snow cover to the land phase of the hydrologic cycle. The rate of water input to the land phase has a dominant effect on the resulting hydrograph. The accumulation of the snow cover, especially in terms of the water-equivalent, is also important because of its influence on the total runoff volume. The areal extent of the snow cover controls the area that can contribute to the runoff process. The remaining snow model components have less effect on the overall simulation results, but are important at certain times on many watersheds.

It should be noted that the variables which define the state of the snow cover in the model (water-equivalent, heat deficit and liquid-water content) are defined in terms of their mean areal value. For example, if an average of 200 MM of snow covers 25 percent of the area, the mean areal water-equivalent is 50 MM. The model keeps an account of both the mean areal value of these snow cover variables and the areal extent of the snow cover.

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Accumulation of the Snow Cover

A critical first step in any snow cover model is to determine whether any new precipitation is in the form of rain or snow. Operationally a variety of information such as upper air data, ground level temperatures, radar returns and visual observations are available to assist the user in determining the form of the precipitation. During calibration a simple criterion, based on the available data, is needed so that the form of the precipitation can be classified correctly most of the time. The snow cover model uses the air temperature to estimate the form of precipitation. The model parameter PXTEMP (DEGC) indicates the temperature which delineates rain from snow. When the air temperature is greater than PXTEMP the precipitation is assumed to be rain and when the air temperature is less than or equal to PXTEMP the precipitation is assumed to be snow.

An incorrect classification of the form of precipitation during certain events may make it impossible or at least difficult to determine a reasonable approximation to the true parameter set for the basin. Random errors in estimating the form of precipitation during events involving small amounts of rain or snow will affect simulation accuracy but should have little effect on the determination of the parameter values. However the improper classification of large amounts of precipitation can have a serious effect on the calibration process. Errors involving large amounts of precipitation are usually easy to detect and should be corrected by adjusting the air temperature data. This is much simpler than obtaining the large quantity of additional data that would be needed to improve the technique used to determine the form of precipitation during the calibration process.

In order to compute the accumulation of the snow cover with reasonable accuracy unbiased estimates of both the form and amount of precipitation are needed. The precipitation data used as input to the snow model are based on point measurements of precipitation from one or more gages. The catch of a precipitation gage can be deficient by a considerable amount during snowfall periods, especially if the gage is not shielded or if the gage is exposed to high winds. Thus it is likely that the precipitation input data are biased estimates of the true precipitation during periods of snowfall. The snow model uses a mean gage catch correction factor to remove this bias. This snowfall correction factor is a model parameter referred to as SCF. Precipitation amounts during individual snow storms can still be in error since SCF is a mean correction factor. However as the number of storms contributing to the snow cover increases, the errors in individual storms will tend to cancel each other.

Precipitation which is classified as snow is adjusted by SCF and added to the existing snow cover. Rain which falls on bare ground immediately enters the soil-moisture accounting model. Rain falling on the snow cover is added to the computed surface melt water.

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Surface Energy Exchange During Non-Melt Periods

The method used to compute snowmelt in the model is described in Section II.2 [Hyperlink]. Energy exchange between the snow cover and the air also occurs during non-melt periods. Non-melt periods are defined as periods when the air temperature is less than MBASE of zero DEGC, whichever is greater. The gain or loss of heat during non-melt periods is assumed to be proportional to the temperature gradient in the upper portion of the snow cover. The snow surface temperature is approximated by the air temperature. The temperature within the snow cover varies as a damped response to changes in the surface temperature. The temperature at some depth below the surface is a function of the recent temperature history at the surface. The model represents the temperature within the snow cover by an antecedent temperature index (ATI, DEGC) which weights the previous surface temperatures by an amount which decreases with time. This index is computed by the equation:

$$ATI_2 = ATI_1 + TIPM * (T_a - ATI_1) \tag{21}$$

where TIPM is the model parameter which determines how much weight is placed on the air temperature for each of the prior periods (range is $0.1 \leq TIPM \leq 1.0$) and the subscripts 1 and 2 refer to the beginning and end of the 6 hour period, respectively

If TIPM equals 1.0, ATI_2 is equal to T_a. If TIPM equals 0.1, the weights assigned to the 6 hour air temperatures are 0.1 for the current period, 0.09 for the previous period, 0.081 for the period before that, etc. Since the antecedent temperature index is representing a snow temperature, the value of ATI cannot exceed zero DEGC. The temperature gradient in the upper portion of the snow cover is thus approximated in the model by the difference between the current air temperature (T_a) and the antecedent temperature index at the beginning of the period (ATI_1).

The gain or loss of heat during non-melt periods can now be expressed as:

$$\Delta D = NM_f * (ATI_1 - T_a) \tag{22}$$

where D is the change in snow cover heat deficit (MMe)
NM_f is the a proportionality factor, referred to as the negative melt factor (MMe/DEGC)

Heat transfer in a snow cover is primarily by conduction. The thermal conductivity of snow is mainly a function of snow density. Since snow density tends to increase as the snow season progresses, the rate of heat transfer should also increase. Equation 22 is an empirical equation, plus snow density is not computed in the model. Thus the seasonal variation of NM_f cannot be directly computed. The model assumes the seasonal variation in NM_f is the same as that for the melt factor, M_f. Therefore, NM_f can be expressed as:

$$NM_f = \left(\frac{M_f}{MFMAX}\right) * NMF \tag{23}$$

where NMF is the maximum negative melt factor (MMe/DEGC/6HR)

NMF is a snow model parameter.

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Snow Cover Heat Storage

The heat storage of the snow cover is represented in the model by the heat deficit. The heat deficit indicates the amount of heat that must be added to return the snow cover to an isothermal state at zero DEGC with the same liquid-water content as when the heat deficit was previously zero. After the heat deficit returns to zero, surface melt water or rainwater can again contribute to liquid-water storage or snow cover outflow. Physically the heat deficit consists of liquid-water that is frozen within the snow cover when heat is lost and snow that is at temperatures below zero DEGC. The physical form of the heat deficit is not important; only the total deficit needs to be known. The model keeps a continuous accounting (on a 6 hour basis) of the total heat deficit.

The total heat deficit of the snow cover can change in three ways. First, the heat deficit can change due to energy exchange during non-melt periods. The change in the heat deficit during such periods is given by Equation 23. Second, the total heat deficit of the snow cover is increased by the heat deficit of snowfall. The heat deficit of the new snow can be computed from Equation 10 by using the specific heat of ice (c_i = 0.5) and by recognizing that the temperature of snowfall cannot exceed zero DEGC. The heat deficit of new snow can be expressed as:

$$D_m = - \frac{P_x * T_p}{160} \qquad (24)$$

where D_m is the heat deficit due to mass changes (snowfall) (MMe)
T_p is the temperature of precipitation (DEGC) (for snowfall assumed equal to the air temperature or zero DEGC, whichever is smaller)

Third, the heat deficit of the snow cover is reduced by the heat released when melt or rainwater freezes within the snow cover. The heat deficit is reduced by the amount of water that freezes. Melt and rainwater will continue to freeze within the snow cover until the total heat deficit reaches zero.

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Areal Extent of Snow Cover

The percent of the area that is covered by snow must be computed to determine the portion of the area over which energy exchange is taking place and, in the case or rain-on-snow, to determine how much rain falls on bare ground. The areal depletion of snow has been shown in several studies to be predominantly a function of how much of the original snow cover remains. The distribution of the snow cover during the accumulation season is influenced by topography, vegetation cover, storm type and wind conditions. Because these factors are stationary or similar from year to year, the snow cover accumulation patterns are also similar. Snowmelt patterns are influenced by topography, vegetation cover and weather conditions and are also similar from year to year. Because of these similarities in snow accumulation and melt patterns, each area should have a reasonably unique areal depletion curve. An areal depletion curve, as used in the model, is a plot of the areal extent of the snow cover versus a ratio which indicates how much of the original snow cover remains. The ratio of the current mean areal water-equivalent to an areal index, A_i (MM) is used. The index value, A_i, is the smaller of: 1) the maximum water-equivalent that has occurred over the area since snow began to accumulate or 2) a preset maximum, SI (MM). The model parameter SI is thus the mean areal water equivalent above which there is always 100 percent areal snow cover. Because of the wide variability in the shape of areal depletion curves from area to area, the depletion curve, as used in the model, is defined by discrete points connected by straight lines. Figure 3 [Bookmark] illustrates what a snow cover depletion curve looks like.

During the depletion of the snow cover, new snow can fall over an area that is partially bare. In this case, the area reverts to 100 percent cover for some period of time, then returns to the areal depletion curve. In the model, the area remains at 100 percent cover until 25 percent of the new snow ablates. Then the area returns, by a straight line, to the point where it was on the depletion curve before the snowfall. This case of new snow falling on a partially bare area, is also illustrated in Figure 3. In reality both the 25 percent figure and the straight line return to the depletion curve seldom occur. However they are reasonable approximations and their effect on the results does not warrant the inclusion of other parameters.

Besides explicitly accounting for the reduction in the areal extent of the snow cover, the depletion curve implicitly accounts for other factors which tend to reduce the amount of energy exchange as the snow cover depletes. The melt factor used in the model represents the melt rate when there is complete areal snow cover. As the snow cover depletes, the portions of the area that remain snow-covered are generally those which melt slowest (e.g., north-facing slopes and heavily forested areas). Thus the relationship between the ratio of the actual melt factor for the snow-covered area to the melt factor based on 100 percent snow cover would be equal to 1.0 when there is 100 percent cover and would decrease in some fashion to a minimum value as the areal extent of the snow cover is reduced. Since this relationship is not explicitly accounted for in the model, it becomes absorbed into the areal depletion curve. The user must bear this in mind when comparing computed and observed values of areal snow cover. This is especially true for low values of areal snow cover because the relationship between melt factors and areal cover will generally exhibit the greatest departure from 1.0 at low values.

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Retention and Transmission of Liquid-Water

Snow crystals retain liquid-water against gravity drainage just like soil particles. In the model the maximum amount of liquid-water that the snow cover can hold is controlled by a model parameter, PLWHC. PLWHC is the percent (decimal) liquid-water holding capacity of the snow. The percent liquid-water is defined in the model as a percentage by weight of the solid (ice) portion of the snow cover. The model assumes PLWHC is a constant for all snow cover conditions. Variations in liquid-water holding capacity with respect to density and grain size have been observed. However any relationships which may exist between liquid-water holding capacity and density or grain size are not well defined.

Liquid-water in excess of that held within the snow cover against gravity drainage is transmitted through the snow cover and becomes snow cover outflow. There is both a delay and damping effect as the excess liquid-water moves through the snow cover. Equations for the transmission of excess liquid-water through the snow cover were developed using data obtained during April and May 1954 from the Central Sierra Snow Laboratory (CSSL) lysimeter. The equations apply to 'ripe' snow (well-aged snow with a spherical crystalline structure which is isothermal at zero DEGC). However in the model these equations are used under all conditions since there is a lack of data and knowledge on the transmission of water through fresh snow. In fresh snow, indications are that both the delay and damping effects of the snow cover on the movement of excess liquid-water are increased. The magnitude of this increase has been difficult to quantify.

In the model the excess liquid-water is first lagged and then attenuated. The equation for lag is:

$$L = 5.33 * \left(1.0 - \exp\left(\frac{-0.03 * WE}{E}\right)\right) \tag{25}$$

where L is the lag in hours
WE is the water-equivalent of the solid (ice) portion of the snow cover (MM)
E is the excess liquid-water (MM/6HR)

The attenuation part of the liquid-water transmission process uses a withdrawal rate which is the portion of the excess liquid-water which drains from storage within the snow cover during a given time interval. The withdrawal rate for a 1 hour time interval as determined from the CSSL lysimeter data is:

$$R_1 = \frac{1.0}{5.0 * \exp\left(\frac{-500 * E_{\ell s}}{WE_s^{1.3}}\right) + 1.0} \tag{26}$$

where R_1 is the 1 hour withdrawal rate
$E_{\ell s}$ is the mean amount of lagged excess liquid-water over the snow-covered area for the current period (IN)
WE_s is the mean water-equivalent of the solid (ice) portion of the snow over the snow-covered area (IN)

The functional forms of Equations 25 and 26 were developed by plotting the experimental data. Final coefficient values were determined by minimizing the squared error between simulated and observed snow cover outflow from the lysimeter. English units are used for E_s and WE_s in Equation 26 because the coefficient values were determined using

English units. There is no simple conversion. Equations 25 and 26 are shown graphically in Figures 4 [Bookmark] and 5 [Bookmark] respectively.

The amount of snow cover outflow on an hourly basis can now be computed as:

$$O_s = (S_1 + E_\ell) * R_1 \tag{27}$$

where O_s is the snow cover outflow (MM/HR)
S_1 is the amount of excess liquid-water in storage in the snow cover at the beginning of the period (MM)
$E\ell$ is the amount of lagged excess liquid-water entering storage during the current period (MM)

The change in the amount of excess liquid-water in storage in the snow cover is computed as:

$$S_2 = S_1 + E_\ell - O_s \tag{28}$$

where S_2 is the amount of excess liquid-water in storage at the end of the period (MM)

The 6 hour snow cover outflow is obtained by merely summing the appropriate hourly outflows.

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Heat Exchange at the Snow-Soil Interface

Heat exchange at the snow-soil interface is usually negligible compared to heat exchange at the air-snow interface on a given day. In some watersheds a small amount of melt takes place continuously at the bottom of the snow cover. Although this melt is small on a daily basis, it can amount to a significant quantity of water over an entire snow season. Ground melt adds to soil moisture storage and helps sustain base flow throughout the winter. The model assumes that a constant amount of melt occurs per day at the snow-soil interface. This constant rate of melt is determined by the parameter DAYGM (units of MM per day). The ground melt is added to the snow cover outflow and to the rain which falls on bare ground to obtain the total rain plus melt. This total rain plus melt becomes the input to the soil moisture accounting model.

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Snow Depth Parameterization

A simplified version of a snow compaction model [Anderson, 1976] was adopted to calculate a snow depth. An extension of the analytical solution by Koren et al. [1999] was derived to calculate the dynamics of the snow density (ρ) due to snow compaction and metamorphism:

$$\rho_{t+dt} = \rho_t \left(\frac{e^{Bw_t} - 1}{BW_t} + e^{A} \right) \tag{29}$$

where

$$B = c_1 * dt * e^{0.08T_s - c_2\rho_t}$$

$$A = c_3 * c_5 * dt * e^{c_4 T_s - 46\beta(\rho_t - \rho_d)}$$

dt is the time step
W_s is a snow water equivalent
T_s is an average snow temperature
c_1, c_2, c_3, c_4, c_5 and ρ_d are constants defined in [Anderson, 1976]:

- c_1 is a fractional increase in density, c_1=0.01, 1/(CM *HR)
- c_2 is a constant estimated by Kojima as 21 CM**3/G
- c_3 is a fractional settling rate at 0 degree temperature for snow densities less than threshold density ρ_d, c3=0.01, 1/HR
- c_4 is a constant, c_4=0.04 1/DEGC
- c_5 is an empirical adjustment factor that accounts for melt metamorphism; best its estimate from limited data is 2.0 (dimensionless)
- ρ_d is a threshold snow density above which a rate of snow metamorphism decreases; a good value is 0.2 G/CM**3

The constant c_5 is applied only if some amount of liquid water are present in snow pack. The constant β=0 if $\rho_t < \rho_d$ and β=1 if $\rho_t > \rho_d$. The first term in Equation 29 accounts for the snow compaction and the second term accounts for metamorphism.

More details on theses parameters can be found in Anderson [1976] and Kojima [1967].

Averaged snow temperature is calculated from an approximate solution of the heat transfer equation using an air temperature change per 6 hourly time interval, $\Delta T_{s,t+\Delta t}$:

$$T_{s,t+\Delta t} = T_{s,t} + \Delta T_{t+\Delta t} \frac{1 - e^{\sqrt{\frac{\pi c}{\lambda 2 \Delta t}} H_s}}{\sqrt{\frac{\pi c}{\lambda 2 \Delta t}} H_s} \qquad (30)$$

where H_s is a snow depth

The snow thermal conductivity, λ, estimated from Djachkova's formulae [Koren, 1991]:

$$\lambda = 0.0442 e^{5.181\rho} \qquad (31)$$

and c is the effective specific volumetric heat capacity of snow:

$$c = c_{ice}\rho + c_{air}(1-\rho-\theta_{liq}) + c_{liq}\,\theta_{liq} \qquad (32)$$

where c_{ice} is the volumetric heat capacity of ice
c_{air} is the volumetric heat capacity of air
c_{liq} is the volumetric heat capacity of water
θ_{liq} is a fraction of liquid water in snow cover

Simulation of snow depth will not effect on results of snow water equivalent simulation.

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PROCESSES NOT EXPLICITLY INCLUDED

The snow model includes representations of those processes which have a significant effect on snow accumulation and ablation provided the effect can be computed with reasonable accuracy using only precipitation and air temperature data. Other physical processes, though possibly significant under certain conditions, are not explicitly included in the model. This is because an adequate quantification of these processes would require additional data. Also, their effect is generally smaller in magnitude, but somewhat similar to the effect produced by a process that is explicitly included in the model. In this case the minor process is implicitly absorbed into the more dominant process. An example is the implicit inclusion of the decrease in the areal melt rate with decreasing snow cover into the areal depletion curve. Besides that example, which has already been discussed, there are several other similar examples of processes which are implicitly absorbed into other model components. These processes are water vapor transfer, interception and re-distribution of snow.

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Water Vapor Transfer

An energy balance method of computing snow cover energy exchange would be needed to accurately estimate the exchange of water vapor between the air and the snow. Formulas to compute vapor transfer require estimates of the wind speed, the vapor pressure of the air and the snow surface temperature. Vapor transfer in the form of condensation is included in Equation 16 for determining snowmelt during rain-on-show periods. During non-rain and non-melt periods vapor transfer is ignored in the model. Either condensation or sublimation is always occurring during these periods. In some watersheds condensation gains tend to balance sublimation losses during the snow season. Thus the net effect on the water balance is negligible. However in other watersheds sublimation losses exceed gains due to condensation. To the extent that this net loss due to vapor transfer is proportional to the water-equivalent of the snow, the loss would be implicitly reflected in the value of the snow correction factor, SCF.

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Interception of Snow

The interception of snow by vegetation and the subsequent loss of part of this intercepted snow is a complex process. During a storm, a portion of the snowfall adheres to the vegetation. The amount of intercepted snow varies according to the type of vegetation, the amount of snowfall and the wind conditions. After the storm, much of the intercepted snow is blown off or falls from the trees and becomes part of the snow cover. Some may melt and drip from the trees to the snow below or run down the tree trunks and some is lost by sublimation or evaporation. Several studies have shown that the total interception loss during the snow season is roughly proportional to the total snowfall. Thus the total snow interception loss would be implicitly reflected in the snow correction factor similar to the net water vapor loss.

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Redistribution of Snow

Snow can be redistributed in a couple of ways. Snowfall can be intercepted by trees and subsequently blown off and redistributed into forest openings. This partly explains why more snow accumulates in forest openings than under the trees. Snow in open areas can be dislodged by the wind from the point where it originally fell. Some of this blowing snow eventually ends up in drifts, while some is lost due to sublimation while in transit or, in extreme cases, some is blown across basin boundaries. The redistribution process is partly responsible for the typical snow accumulation pattern of a given area. This aspect of the redistribution process is reflected in the snow cover areal depletion curve. Losses due to blowing snow are, to some extent, implicitly reflected in the snow correction factor.

The effect of blowing snow can be more severe at a point than within a basin. Even though snow is redistributed by the wind, it generally stays within the basin where it fell. However the snow cover at a point can experience significant losses or gains due to blowing snow. The effect of blowing snow on the value of SCF at a point, both in terms of the mean value and the variability, can exceed the effect of the precipitation gage catch deficiency. Thus it should be noted that because of blowing snow plus net vapor transfer and interception losses, SCF can be less than 1.0. This could occur for an area, but is more likely at a point.

The redistribution of snow may also result in portions of an area being blown free of snow. In some places this is a frequent occurrence. The model always assumes that the entire area is snow covered after any accumulation period. The best approximation possible for an area in which large portions are generally blown free of snow would be obtained by using an areal depletion curve that drops off very rapidly when ablation begins.

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APPLICABILITY OF THE SNOW MODEL

The model can be used to represent the snow accumulation and ablation process at a point or over an area. The area can be an entire headwater basin or a local area or a basin can be subdivided on the basis of elevation or such physiographic factors as aspect or vegetation cover. When there is a considerable difference in the quantity of snow and the timing of melt over a basin, it is absolutely necessary to subdivide the basin into 2 or 3 parts in order to obtain adequate simulation results. In such basins subdividing into two subareas is usually adequate. The use of two subareas has proven to be adequate in basins in the western United States with elevation ranges up to 2500 M (8200 FT). If a watershed is subdivided into too many parts, it becomes difficult to determine reasonably unique parameter values solely from a single output hydrograph. Although the snow model parameters are related to climatic and physiographic factors these relationships are not precise. Thus calibration of the model to a basin that has been overly subdivided becomes a curve-fitting exercise unless there is plenty of supporting data, such as snow course or additional streamflow data, on which to base the parameter values for each subarea.

The model gives the best results when the day to day and year to year variability in meteorological conditions affecting snow cover accumulation and energy exchange are small. The meteorological conditions affecting accumulation are those which influence the spatial distribution of precipitation, the average gage catch deficiency during snowfall, the net vapor transfer and the re-distribution of snow. Air temperature is generally a good index to energy exchange even with a certain amount of variability in meteorological conditions. Situations where air temperature is not a good index to snow cover energy exchange include:

- clear, cool periods after the snow has ripened
- windy periods with dew points well above zero DEGC
- abnormally warm, but calm periods

In the first two cases the use of a temperature index typically results in an underestimation of snowmelt, whereas in the third case, melt is overestimated. Since the model does not explicitly account for variability in meteorological conditions affecting snow cover accumulation and energy exchange other than through variations in precipitation amount and air temperature, fluctuations in meteorological conditions are likely to cause errors. The simulation errors will be larger in locations where there is a great deal of variability in meteorological conditions.

Since most of the model parameters are indices to snow cover accumulation and energy exchange, the model must be calibrated against observed data. At a point, the calibration is typically based on a comparison of observed and computed water-equivalent. For areal applications, the model is used in conjunction with a soil moisture accounting and a channel system model. The calibration model is useful for river forecasting and certain types of water resource planning and design studies. However since the relationships between the model parameters and various climatic and physiographic characteristics are not precisely defined, the model is not as well suited for use in ungaged areas or for predicting the effect of land use changes on snow accumulation, melt rates and streamflow. Also, it is difficult to predict, without recalibrating the model, the true effect on model parameters of changes in land use or physiographic conditions, such as the changes resulting from a large forest fire.

The snow model has been applied to many areas of the United States. These include areas with a variety of climatic and physiographic conditions such as New England, the Upper Midwest, the Rocky Mountains, the Sierra Nevada and Alaska. The results have typically been good in all these areas, as long as the watershed is properly subdivided, the form of precipitation is generally correct, the input data are reasonably unbiased estimates of the true input and the model is properly calibrated.

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SUMMARY OF SNOW MODEL PARAMETERS

There are six major parameters in the snow cover model. This includes the areal depletion curve which is not a single-valued parameter, but a series of numbers. These are the parameters which typically have the greatest effect on the simulation results. Thus most of the effort during calibration should be devoted to determining the proper value of these parameters. The major parameters in the snow model are:

1. SCF - A multiplying factor which adjusts precipitation data for gage catch deficiencies during periods of snowfall and implicitly accounts for net vapor transfer and interception losses. At a point, SCF also implicitly accounts for gains or losses due to drifting.

2. MFMAX - Maximum melt factor during non-rain periods, assumed to occur on June 21 (MM/DEGC/6HR).

3. MFMIN - Minimum melt factor during non-rain periods, assumed to occur on December 21 (MM/DEGC/6HR).

4. UADJ - The average wind function during rain-on-snow periods (MM/MB).

5. SI - The mean areal water-equivalent above which there is always 100 percent areal snow cover (MM).

6. Areal depletion curve - Curve which defines the areal extent of the snow cover as a function of how much of the original snow cover remains. Also implicitly accounts for the reduction in the melt rate that occurs with a decrease in the areal extent of the snow cover.

There are six minor parameters in the snow model. These are parameters which normally can be determined in advance on the basis of a knowledge of the typical climatic and snow cover conditions for the area. These initial estimates may require one or two slight adjustments, but the determination of the appropriate values for these parameters should always be a small part of the calibration process. The minor parameters in the snow model are:

1. NMF - Maximum negative melt factor (MMe/DEGC/6HR)

2. TIPM - Antecedent temperature index parameter (range is $0.1 \leq TIPM \leq 1.0$)

3. PXTEMP - The temperature which delineates rain from snow (DEGC)

4. MBASE - Base temperature for snowmelt computations during non-rain periods (DEGC)

5. PLWHC - Percent (decimal) liquid water holding capacity, indicates the maximum amount of liquid water that can be held against gravity drainage in the snow cover

6. DAYGM - Constant amount of melt which occurs at the snow-soil interface whenever snow is present (MM)

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Evapotranspiration Reduction Parameter

There is another parameter which should be mentioned in this section. This is the effective forest cover parameter, EFC. The parameter EFC is used in the soil-moisture accounting model, but since it is used only when snow is on the ground, it is commonly associated with the snow model. When the area is completely covered by snow, evapotranspiration can occur only from the portion of the area covered by active vegetation. This usually corresponds to the portion of the area covered by conifers, which are generally the only active vegetation during the snow season. The parameter EFC designates the decimal fraction of the area over which evapotranspiration occurs when there is 100 percent areal snow cover. The evapotranspiration demand when snow is on the ground can be expressed as:

$$ET_s = (EFC + (1.0 - EFC) * (1.0 - A_s)) * ET \qquad (29)$$

where ET_s is the watershed evapotranspiration demand when snow is on the ground (MM)
A_s is the areal extent of snow cover (decimal fraction)
ET is the watershed evapotranspiration demand when the ground is bare of snow (MM)

The parameter EFC is similar to the minor snow model parameters in that it can usually be estimated from a knowledge of the forest cover of the area and is seldom changed during the calibration process.

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Figure 1. SNOW-17 physical processes

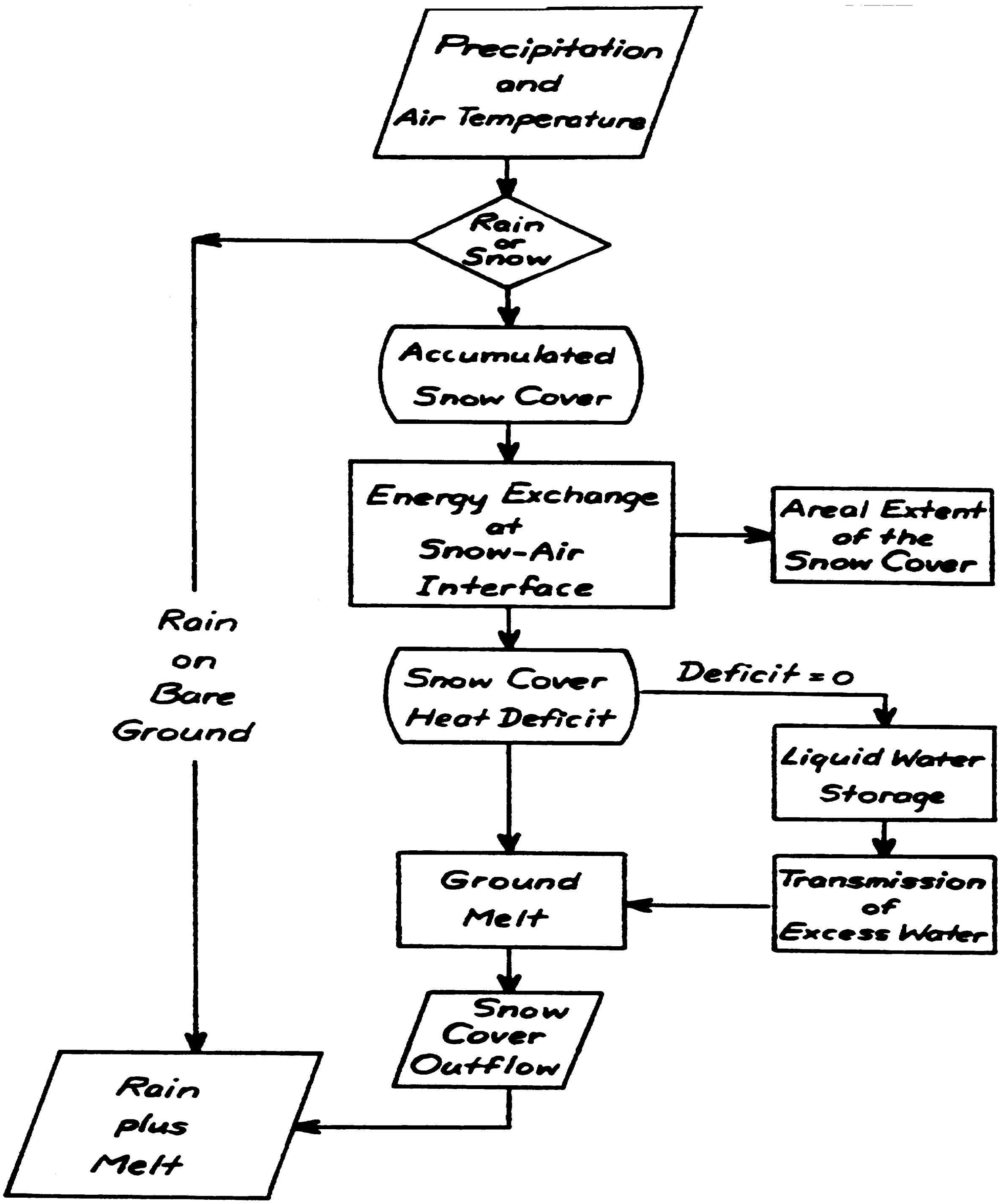

Figure 1. SNOW-17 Physical Processes

Figure 2. Seasonal variation in melt factors used during non-rain periods

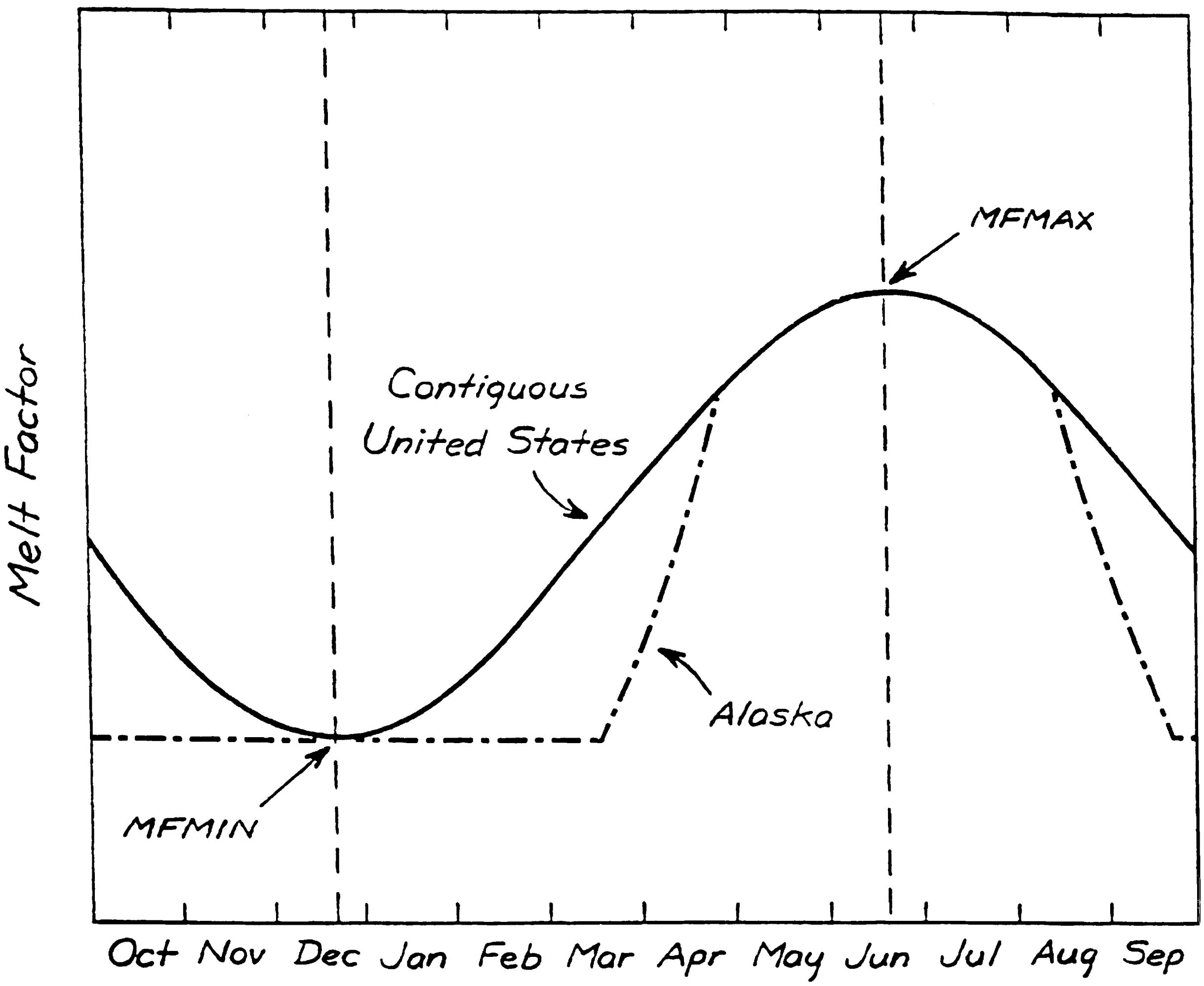

Figure 2. Seasonal variation in melt factors used during non-rain periods

Figure 3. Sample snow cover areal depletion curve

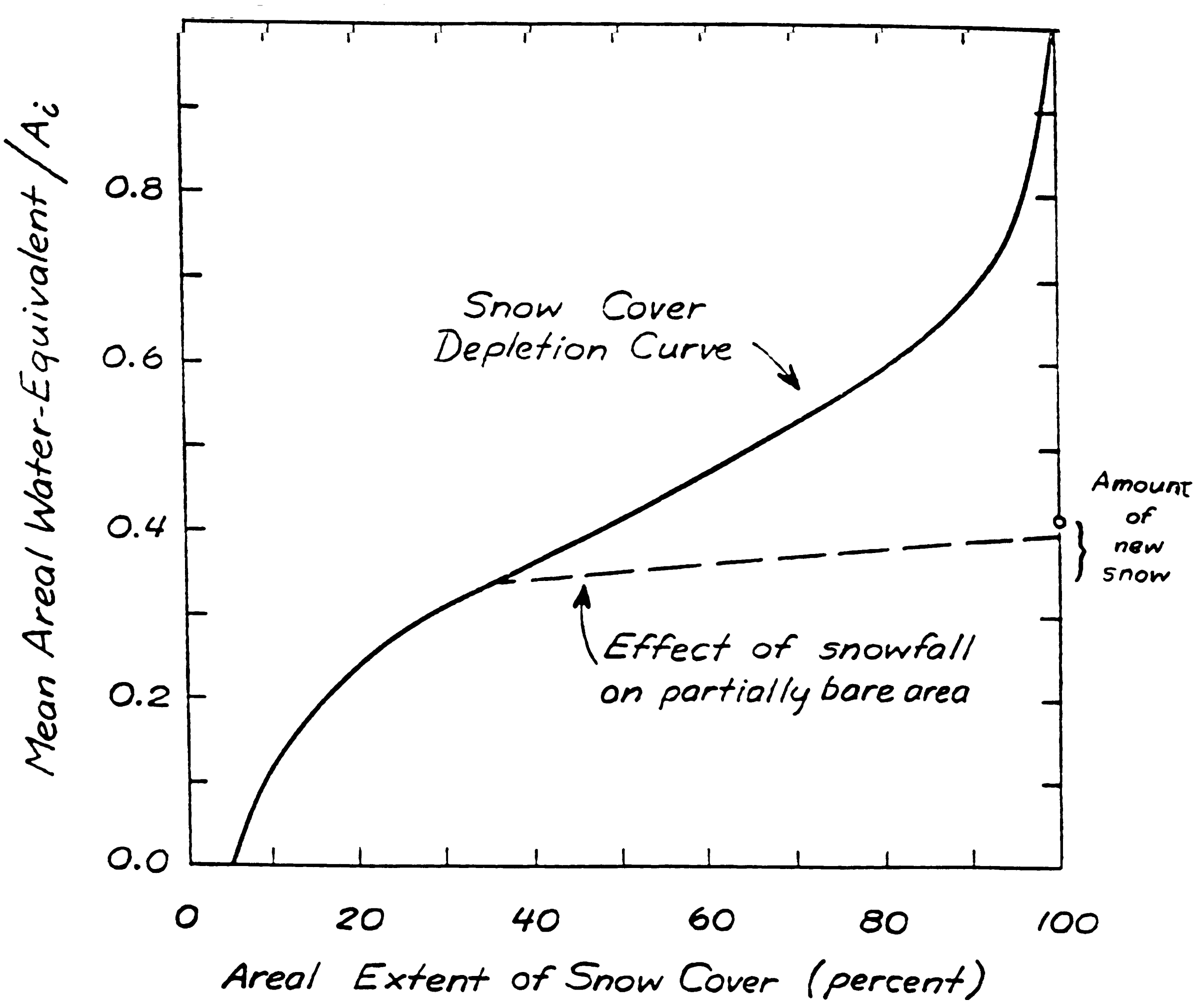

Figure 3. Sample snow cover areal depletion curve.

Figure 4. Lag applied to excess water moving through a snow cover

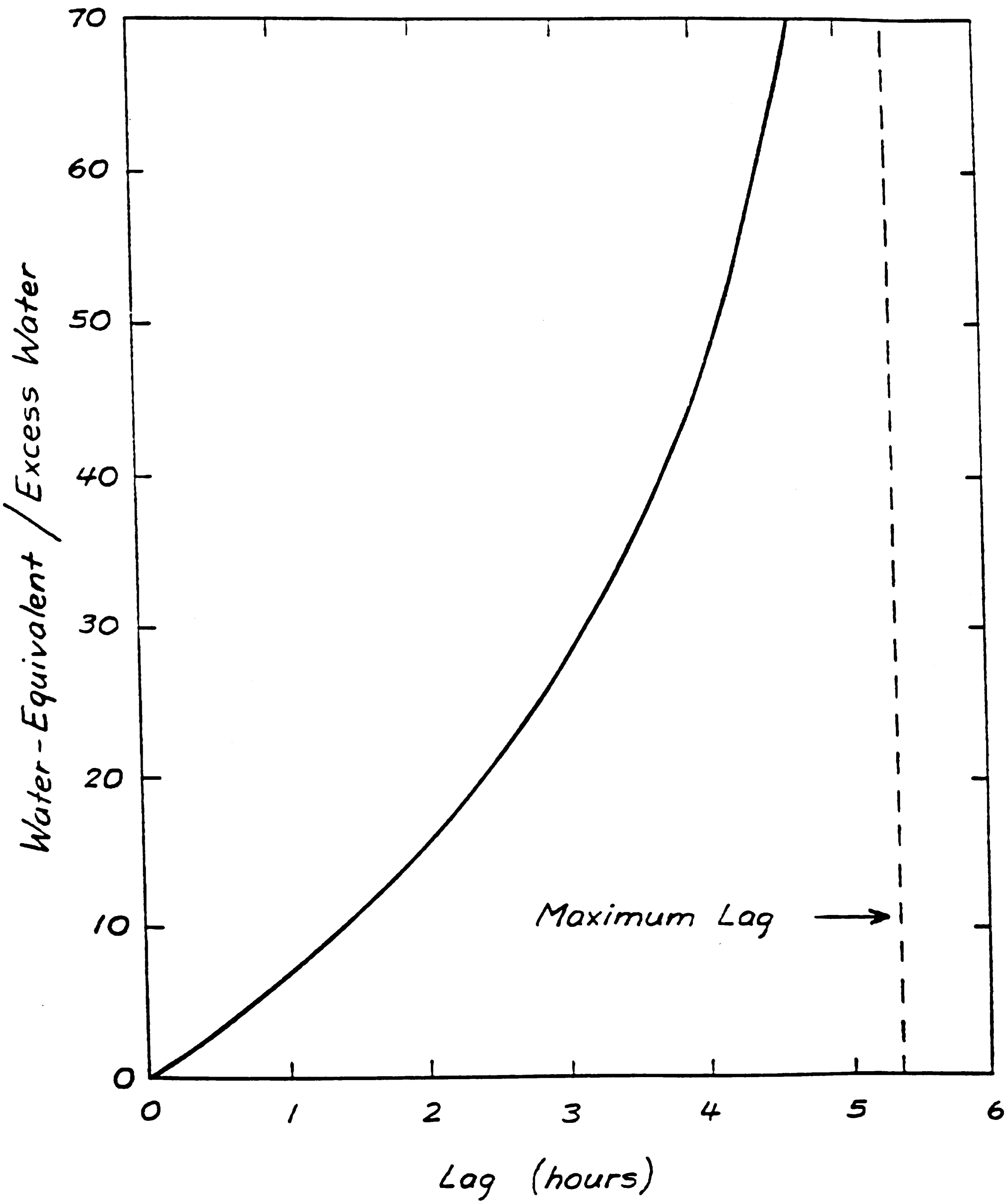

Figure 4. **Lag applied to excess water moving through a snow cover.**

Figure 5. Attenuation of excess water moving through a snow cover

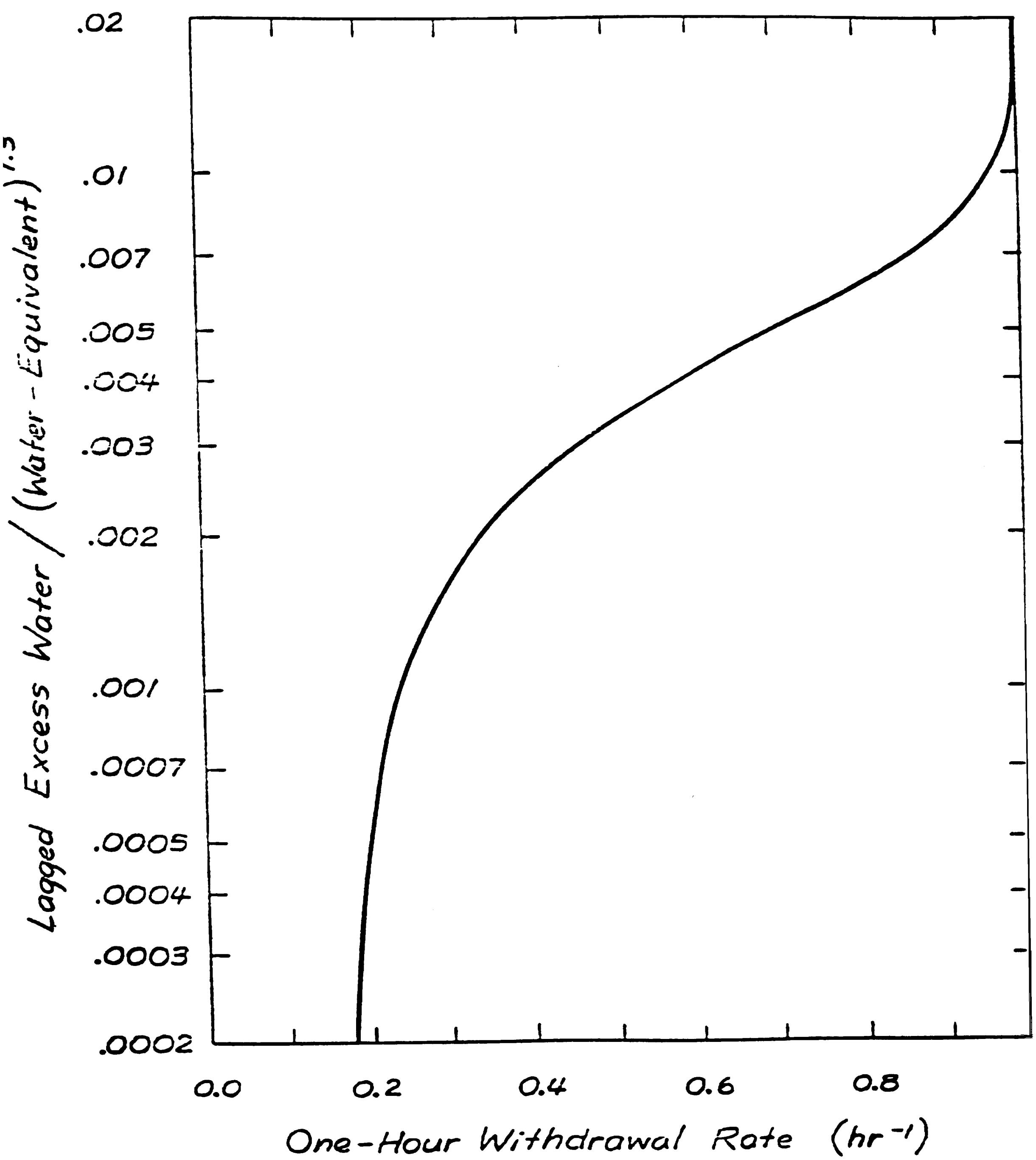

Figure 5. **Attenuation of excess water moving through a snow cover.**